7.2.4 Seismic velocity, attenuation and rock properties

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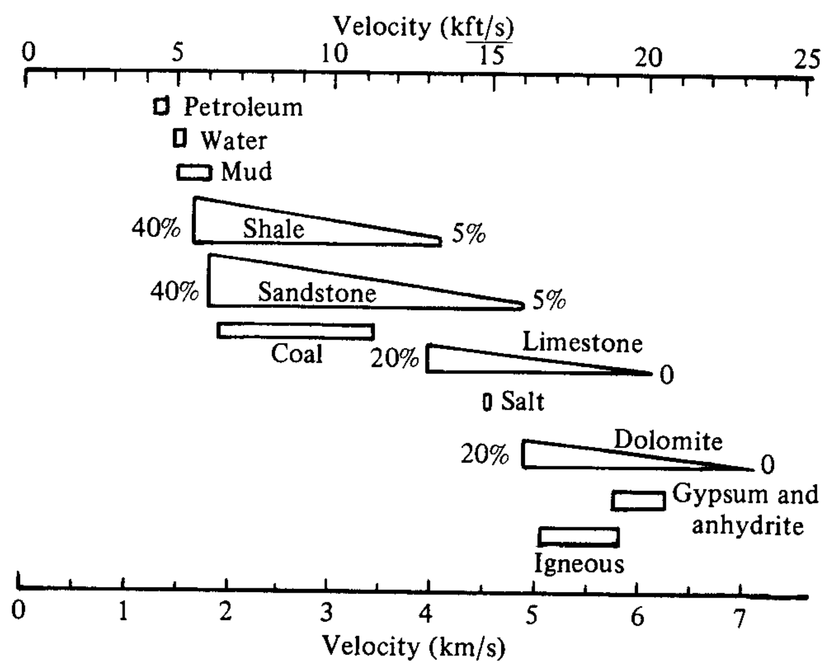
- Attenuation

Seismic surveys yield maps of the distribution of seismic velocities, interfaces between rock units and, ideally, of reflection coefficients at these interfaces. The velocities of crustal rocks vary widely as the following figure shows.



Generally, the velocities depend on the elastic modulii and density via:

$$\boldsymbol{V_P} = \sqrt{\frac{\boldsymbol{K} + \frac{4}{3}\mu}{\rho}} \qquad \text{and} \qquad \boldsymbol{V_S} = \sqrt{\frac{\mu}{\rho}}$$

These elastic constants, and densities, in turn depend on the properties that the geologist or engineer use to characterize the rock such as porosity, fluid saturation, texture etc. A review of the relationships between the intrinsic rock properties and the measured velocities or reflectivities is needed before seismic survey results can be interpreted quantitatively in terms of lithology. Many of these relationships are empirical – velocities are found to be related to certain rock units in a given locale by actual laboratory measurements on core samples of the rock or soil.

It is observed from seismic surveys that velocities generally increase with depth. Densities also increase with depth so it must be that the bulk and shear modulii increase faster than the density. In seismic exploration there are many empirical relationships between velocity and depth of burial and geologic age.

The relationship between intrinsic rock properties such as porosity, fracture content, fluid content and density and velocity underlie the empirical relationships mentioned above.

Rock properties that affect seismic velocity

1) Porosity.

A very rough rule due to Wyllie is the so called time average relationship:

$$\frac{1}{V_{bulk}} = \frac{\phi}{V_{fluid}} + \frac{1-\phi}{V_{matrix}}$$

where ϕ is the porosity.

This is not based on any convincing theory but is roughly right when the effective pressure is high and the rock is fully saturated.

2) Lithification.

Also known as cementation. The degree to which grains in a sedimentary rock are cemented together by post depositional, usually chemical, processes, has a strong effect on the modulii. By filling pore space with minerals of higher density than the fluid it replaces the bulk density is also increased. The combination of porosity reduction and lithification causes the observed increase of velocity with depth of burial and age.

3) Pressure.

Compressional wave velocity is strongly dependant on effective stress. [For a rock buried in the earth the confining pressure is the pressure of the overlying rock column, the pore water pressure may be the hydrostatic pressure if there is connected porosity to the surface or it may be greater or less than hydrostatic. The effective pressure is the difference between the confining and pore pressure.]

In general velocity rises with increasing confining pressure and then levels off to a “terminal velocity” when the effective pressure is high. The effect is probably due to crack closure. At low effective pressure cracks are open and easily closed with an increase in stress (large strain for low increase in stress—small K and low velocity). As the effective pressure increases the cracks are all closed, K goes up and the velocity increases.

Finally even at depth, as the pore pressure increases above hydrostatic, the effective pressure decreases as does the velocity. Overpressured zones can be detected in a sedimentary sequence by their anomalously low velocities.

4) Fluid saturation.

From theoretical and empirical studies it is found that the compressional wave velocity decreases with decreasing fluid saturation. As the fraction of gas in the pores increases, K and hence velocity decreases. Less intuitive is the fact that V_s also decreases with an increase in gas content. The reflection coefficient is strongly affected if one of the contacting media is gas saturated because the impedance is lowered by both the density and velocity decreases.

Velocity in unconsolidated near surface soils (the weathered layer)

The effects of high porosity, less than 100% water saturation, lack of cementation, low effective pressure and the low bulk modulus (due to the ease with which native minerals can be rearranged under stress) combine to yield very low compressional and shear wave velocities in the weathered layer. V_p can be as low as 200 m/sec in the unsaturated zone (vadose zone) – less that the velocity of sound in air!

Attenuation

It is observed that seismic waves decrease in amplitude due to spherical spreading and due to mechanical or other loss mechanisms in the rock units that the wave passes through.

The attenuation for a sinusoidal propagating wave is defined formally as the energy loss per cycle (wave length) Δ E/E where E is the energy content of the wave.

Mathematically, the propagating wave $A = A_0 e^{i\omega t - ikx}$, get an added damping term $e^{-\alpha x}$ so the solution becomes

$$A = A_0 e^{i\omega t - ikx} e^{-\alpha x}$$

[We can apply this to the definition of attenuation Δ E/E by substituting A^2 for the energy at two points at distance λ (the wavelength) apart and we find $\frac{\Delta E}{E} = 2\alpha\lambda$.]

There are many theories for explaining attenuation in rocks. Friction, included by including a velocity term in the governing differential equation for the displacement does not explain laboratory measurement. Various other damping mechanisms such as viscous flow (Biot Theory) have some success but much important work remains to be done in this area (especially for unconsolidated material where the attenuation is very high). Some of the theories predict attenuation as well as dispersion (the variation of velocity with frequency).

Experimentally it is found that the attenuation coefficient α depends on frequency and that there is little dispersion. In fact to a good approximation attenuation can be described by $A = A_0 e^{-\beta f x}$. With x in meters and f in Hertz, a typical shale has a $\beta = 10^{-4}$. So at one Hertz the amplitude falls to A_0/e at 10 km. But at 1000 Hz it falls to A_0/e in 10 m. The attenuation may be as much as 10 times greater in unconsolidated sediments.

Another important attenuation mechanism is the reduction in amplitude of a wave by the scattering of its energy by diffraction by objects whose dimensions are on the order of the wavelength. If *a* is an average linear dimension of velocity inhomogeneities then the attenuation coefficient is given approximately by:

$$\alpha \approx \frac{\boldsymbol{a}^3}{\lambda^4}$$

So attenuation increases rapidly with decreasing wavelength. Consider attenuation is an unconsolidated medium with a velocity of 250 m/sec and a frequency of 1000 Hz. Then, $\lambda = 0.25$ m, and $\alpha = a^3 \times 256$. The wave would fall to 1/e of its initial amplitude when a = 157 m.

It might be reasonable to expect inhomogeneities with a characteristic dimension on the order of 15 cm in the overburden so it is likely that the very high attenuation observed in near surface unconsolidated sediments is due to scattering.